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Optically Stimulated Luminescence Dating Supports Central Arctic Ocean CM-scale Sedimentation Rates

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[1] This paper presents new results from Optically Stimulated Luminescence (OSL) dating on a sediment core raised from the crest of the Lomonosov Ridge in the central Arctic Ocean. There has been much debate about dating sediment cores from the central Arctic Ocean and by using an independent absolute dating technique we aim to test whether or not relatively fast, cm-scale/ka, sedimentation rates were typical of Arctic's Pleistocene depositional mode. On the basis of mainly paleomagnetic reversal stratigraphy, many previous studies suggest mm-scale/ka sedimentation rates. A common feature in these studies is that the first down core paleomagnetic negative inclination is consistently interpreted as the Brunhes/Matuyama boundary at about 780 ka. Our OSL dating results indicate that this assumption is not generally valid, and that the first encountered negative inclination represents younger age excursions within the Brunhes Chron, implying reinterpretation of many published core studies where paleoenvironmental reconstructions have been made for the central Arctic Ocean. Our dating results furthermore corroborates a correlation of the uppermost 2–3 m of the Lomonosov Ridge cores to a well-dated core located off the Barents-Kara Sea margin that in turn is correlated to cores in the Fram Strait. Valuable information on the paleoceanographical evolution in the Arctic Ocean from MIS 6 to the Holocene is given through this correlation of records from the central Arctic Ocean to records off the Eurasian continental margin.

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1. Introduction

[2] The Arctic Ocean is the smallest of the World Ocean's, but it is nevertheless considered to exert a considerable influence on global climate [e.g., *Alley*, 1995; *Driscoll and Haug*, 1998; *COSOD I*, 1981; *COSOD II*, 1987; *COMPLEX*, 1999]. Sediments preserved on Arctic's seafloor contain a record of the paleoenvironmental variability of the northern polar ocean. The permanent Arctic Ocean sea-ice cover has prevented samples to be acquired from this extreme environment using conventional research ship operations. But modern icebreakers have been able to penetrate the pack-ice during the last 15 years, to collect numerous short sediment cores from the central Arctic Ocean [e.g., *Thiede*, 1988; *Fütterer*, 1992; *Grantz*, 1993; *Jokat*, 1999]. There has been much debate about the dating of these cores as well as of the dating of earlier short cores retrieved from ice islands [e.g., *Herman*, 1974; *Sejrup*, 1984; *Macko and Aksu*, 1986; *Thiede et al.*, 1990; *Frederichs*, 1995; *Darby et al.*, 1989, 1997; *Jakobsson et al.*, 2000]. It is only within the framework of an accurate chronology that the paleoenvironmental history, as preserved in Arctic sediment cores, can be deciphered.

[3] Sporadic occurrences or absence of calcareous and siliceous microfossils have prevented the use of conventional biostratigraphic and isotopic dating methods. Therefore, most age models of existing central Arctic Ocean sediment cores largely have relied on the interpretation of paleomagnetic reversal stratigraphy [e.g., *Steuerwald et al.*, 1968; *Clark*, 1970; *Clark et al.*, 1980; *Aksu*, 1985; *Aksu et al.*, 1988; *Witte and Kent*, 1988; *Poore et al.*, 1994; *Frederichs*, 1995; *Schneider et al.*, 1996; *Nowaczyk et al.*, 2001]. Measured paleomagnetic inclination changes may be interpreted either as short excursions of the Earth's magnetic field, on the order of 5–10 ka, or full magnetic reversals, which are followed by stable polarity directions lasting several hundred thousand years [*Gubbins*, 1999]. Independent dating methods must be employed to guide the interpretation of the polarity pattern otherwise the alternating sequence of polarity directions is open to conflicting interpretations [e.g., *Frederichs*, 1995], resulting in estimates of sedimentation rates that may easily differ by an order of magnitude.

[4] Arguments for relatively fast sedimentation rates (on the order of cm/ka) from Arctic Ocean sediment cores have been raised by a few studies using different dating methods: for example, amino acid epimerization dating of cores retrieved from the Amerasia Basin [*Sejrup et al.*, 1984], calcareous nannofossil stratigraphy in cores from the Lomonosov Ridge and Eurasian basins and ridges [e.g., *Gard*, 1993], and paleointensity studies of cores from the Makarov Basin [e.g., *Nowaczyk et al.*, 2001]. Yet most previous studies of Arctic Ocean sedimentation rates argue for slow sedimentation rates (on the order of mm/ka) [e.g., *Clark et al.*, 1980; *Aksu*, 1985; *Aksu et al.*, 1988; *Witte and Kent*, 1988; *Poore et al.*, 1994; *Clark et al.*, 2000]. Studies suggesting mm-scale/ka sedimentation rates have one significant common feature: they interpret the first down core paleomagnetic negative inclination as the Brunhes/Matuyama boundary, presently dated at 780 ka [*Shackleton et al.*, 1990].

[5] *Jakobsson et al.* [2000] published an age model from a sediment core retrieved from the Lomonosov Ridge in the central Arctic Ocean that was based on the hypothesis that manganese enriched, brown colored sediment units represent warm phases, that is, interglacials (core location, Figure 1). Calcareous nannofossil biostratigraphy, using presence of *Emiliania huxleyi*, provided a crucial biochronologic control point, indicating that the measured negative inclinations in this core represent excursions rather than reversals. These data allowed the alternating sequence of manganese rich/poor sediment units to be correlated to low latitude $\delta^{18}O$ variations. Using this approach, Pleistocene sedimentation rates derived from the Lomonosov Ridge core 96/12-1pc were 2–3 cm/ka (average 2.8 cm/ka) in sediments younger than Marine Isotope Stage (MIS) 5.1 at ca 80 ka, when a drastic change occurred (Figure 2). Rates prior to 80 ka show values ranging between 0.2 and 1.6 cm/ka, with an average of 0.5 cm/ka. In contrast to these cm-scale sedimentation rates, where a 7 + m long core represents ~1 Ma, results from previous studies of cores from the Lomonosov Ridge

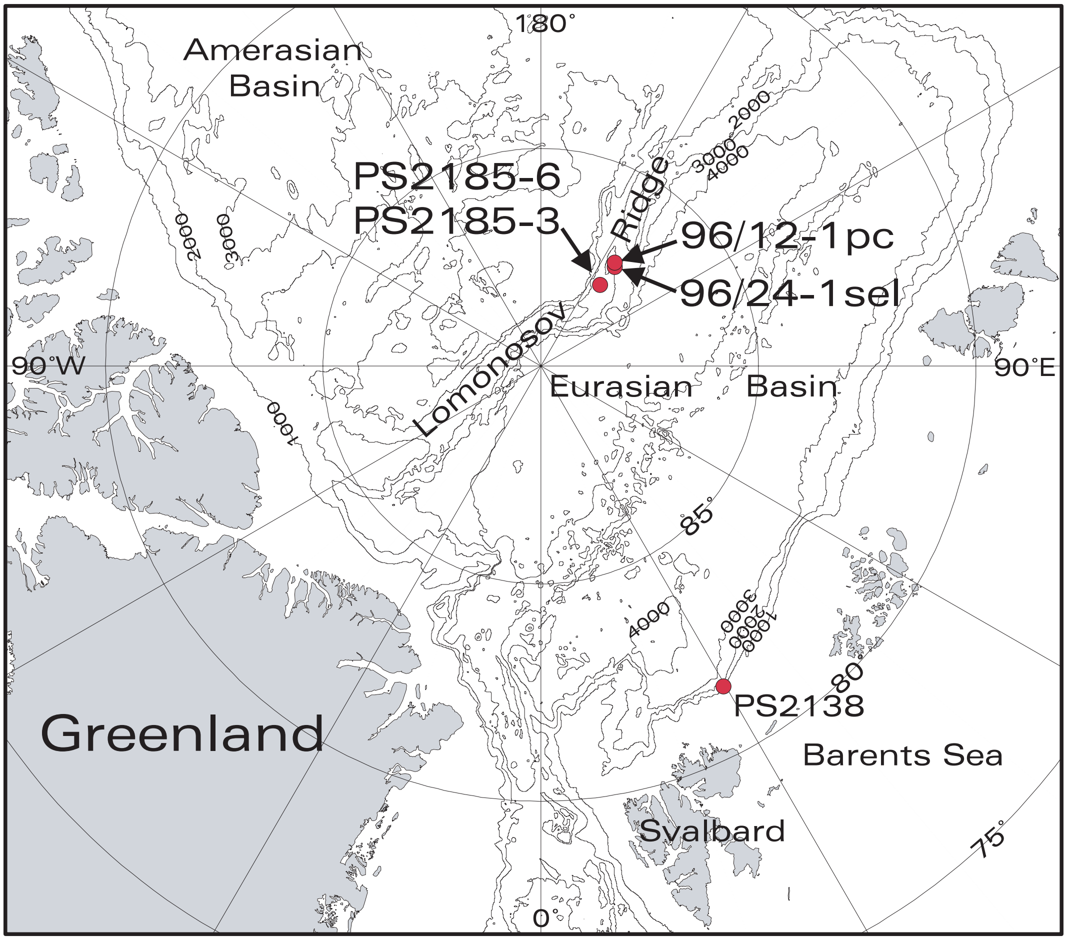

Figure 1. Map showing the locations of sediment cores from the Arctic Ocean discussed in this study.

suggest sedimentation rates on the order of mm per thousand years, where a 7+ m long core represents ~5 Ma [e.g., *Morris et al.*, 1985].

[6] Here we present new results from Optically Stimulated Luminescence (OSL) dating on a core that can be correlated unambiguously to other critical cores from the Lomonosov Ridge. The main objective for this study is to further test, using an independent absolute dating technique, whether relatively fast, cm-scale/ka, sedimentation rates were typical of the Pleistocene depositional mode in the central Arctic Ocean. A verification of this hypothesis would imply reinterpretation of most published core studies, in which paleoenvironmental reconstructions have been made from chronologies relying on an inferred Brunhes/Matuyama boundary at the first down core negative polarity interval. Our dating results are furthermore supported by correlations of the uppermost 2–3 m of the Lomonosov Ridge cores to a well-dated core located off the Barents-Kara Sea margin that in turn is correlated to cores in the Fram Strait. This links paleoenvironmental records from the central Arctic Ocean to records off the Eurasian continental margin and gives valuable information on the paleoceanographical evolution in the Arctic Ocean from MIS 6 at ca 150 ka to the Holocene.

2. Testing Arctic Ocean Age Models Using OSL Dating

[7] Luminescence dating offers an independent absolute age control for clastic sediments [*Murray and Olley*, 1999, 2002]. This technique uses the build up of a trapped electron population in natural minerals, such as quartz, as a chronometer. The electron population is initially set to zero by daylight exposure during transport (by wind, water, or ice), and increases with time because of exposure

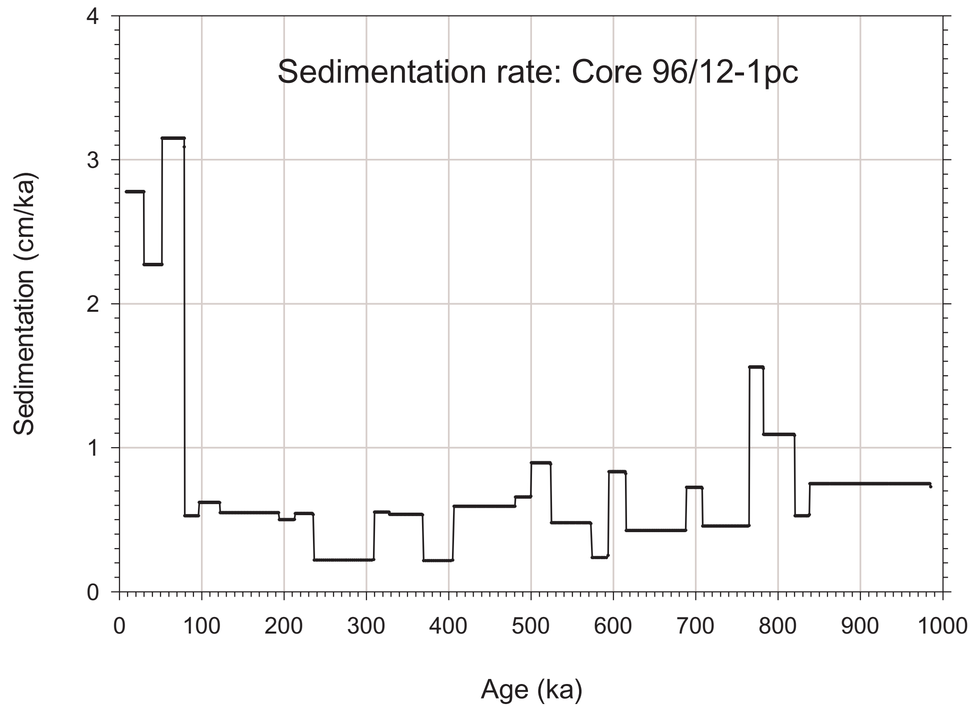

Figure 2. Variation in sedimentation rate for core 96/12-1PC as derived from the chronology by *Jakobsson et al.* [2000].

to naturally occurring radiation from the sediments. OSL dating is widely used in terrestrial deposits, and provides much of our present understanding of the timing of the last glacial cycle in Arctic Eurasia [e.g., *Alexanderson et al.*, 2001; *Houmark-Nielsen et al.*, 2001; *Mangerud et al.*, 2001]. The dating technique is usually applied to sediments that are younger than ca 150 ka (depending on the dose rate). It has not been widely applied in marine cores, primarily because accurate and quick isotopic methods (^{18}O and ^{14}C) have been available for cores raised from equatorial through subpolar regions. But recent work on marine material of Eemian age from Denmark has proved very encouraging, yielding a mean age of 119 ± 6 ka for 22 samples [*Murray and Funder*, 2003].

[8] Sediment transport mechanisms play important roles in the application of luminescence dating because of the importance of light exposure to the resetting of the OSL signal. In the Arctic Ocean, clastic sediments are chiefly transported to its central parts by sea-ice or icebergs [*Clark and Hanson*, 1983; *Hebbeln and Wefer*, 1991]. Sea-ice transportation dominates today [*Eicken et al.*, 2000], which was probably true also for previous interglacials. In contrast, iceberg transportation increased substantially during glacials [*Elverhøi*, 1998]. In the case of rafting via sea-ice, the majority of sediment particles are likely to be sufficiently exposed to daylight in order to be set to zero. The contrary is the case for rafting via icebergs, because large amounts of the transported sediment particles will be carried at a substantial water depth incorporated into the glacier ice, and many of these particles may also have been derived from subglacial erosion, lacking a Pleistocene history of daylight exposure. It follows that the manganese enriched, brown colored sediment units containing microfossils, interpreted to represent interglacial conditions [*Phillips and Grantz*, 1997; *Jakobsson et al.*, 2000], are likely to host well-bleached clastic sediments due to a dominant transport via sea-ice. These sediment units, therefore, should be favorable for OSL dating. On the other hand, the more coarse grained units deposited during glacial conditions [*Jakobsson et al.*, 2001] should not be suitable for OSL dating. It is thus hypothesized that incomplete initial zeroing of iceberg transported particles, or particles derived

from subglacial shelf erosion, could yield age inversions in the stratigraphic record, with anomalously old OSL ages in glacial units in comparison to OSL ages derived from underlying interglacial units.

[9] Although the measurement of luminescence signals from single sand-sized quartz grains has been suggested as an approach to identifying incompletely bleached sediments [e.g., *Murray and Olley*, 1999], this is unlikely to be useful in this case. This is because this technique relies on the identification of heterogeneous bleaching of the grains in a deposit, whereas in our model, the iceberg transported grains should not have received any light exposure. However, such single grain techniques might be of value in distinguishing between grains transported by relatively transparent sea-ice and by icebergs.

3. Methods

[10] Samples for OSL dating were extracted from sediment core 96/24-1sel. This core was raised from the Lomonosov Ridge at 87.183° latitude and 144.606° longitude (water depth 980 m) during the *Arctic Ocean 96* expedition using the Swedish icebreaker *Oden* (Figure 1). The 400 cm long and 10 cm diameter core was contained in a black, nontransparent plastic core liner before it was split and sampled for OSL dating in a dark room at the Nordic Laboratory for Luminescence Dating (Risø, Denmark). Eleven samples have been examined, 9 sampled every 5 cm from between 170 and 215 cm, and 2 between 65 and 75 cm. The core had partially dried out before sampling, and this made the water content estimations unreliable. The water content is important because it attenuates the dose rate; the larger the water content the smaller the dose rate, and so the older the age - increasing the water content by 1% increases the age by about the same amount. We have used water contents measured on sediments from an adjacent core 96/12-1pc, for which the correlations with the present samples are unambiguous (see below, and Figure 3).

[11] All radionuclide concentrations were determined using high-resolution gamma spectrometry, following the methods described by *Murray et al.* [1987]. Surficial deep-sea sediments are well known to contain an excess of ^{230}Th over the series parent ^{238}U; this excess decays with depth, and so introduces a time dependent component into the dose rate. However in our samples the average $^{226}Ra/^{238}U$ ratio is 0.89 ± 0.07 (n = 9; not including sample 006612 at 175 to 180 cm); it can be assumed that ^{226}Ra is at, or close to, equilibrium with its parent ^{230}Th at these depths. Mean ^{238}U, ^{226}Ra, ^{232}Th and ^{40}K concentrations are 30.8 ± 1.4, 26.9 ± 1.8, 45.9 ± 2.0 and 740 ± 30 $Bq.kg^{-1}$, respectively. Thus the $^{226}Ra/^{238}U$ ratio (and thus the $^{230}Th/^{238}U$ ratio) is close to, or slightly below, unity after only 1 to 2 ^{230}Th half-lives. This suggests that sediment delivery was sufficiently high at the time these layers were deposited on the ocean floor that precipitating ^{230}Th did not contribute significantly to the total ^{230}Th sediment budget. The remaining sample (175 to 180 cm) is unusual, in that the $^{226}Ra/^{238}U$ ratio is much higher, at 2.4 ± 0.4, although the ^{238}U, ^{232}Th and ^{40}K concentrations are very similar to the means given above. We are forced to conclude that the sedimentation rate must have decreased for a period during the deposition of this layer, such that precipitation of ^{230}Th from the water column was able to contribute significantly to the total ^{230}Th concentration in the depositing sediments at that time. From a dosimetric point of view, this means that only the sample at 175 to 180 cm may require any allowance for a time dependent change in dose rate, but in any event the effect is <8%, because the majority of the gamma dose comes from the surrounding layers, and ^{230}Th and daughters only contribute <10% of the total beta dose in this sample. The derived dose rates are summarized in Table 1.

[12] The radiation dose absorbed by the sample since last exposure to daylight (the equivalent dose, D_e) was measured using the single-aliquot regenerative-dose (SAR) protocol [*Murray and Wintle*, 2000], and these doses are also summarized in Table 1, together with the number of aliquots (n) used to derive the mean and standard error. The doses are all >250 Gy; this is in the region of the quartz growth curve where significant nonlinearity is observed. From an examination of the growth curves for individual aliquots of these samples, it is

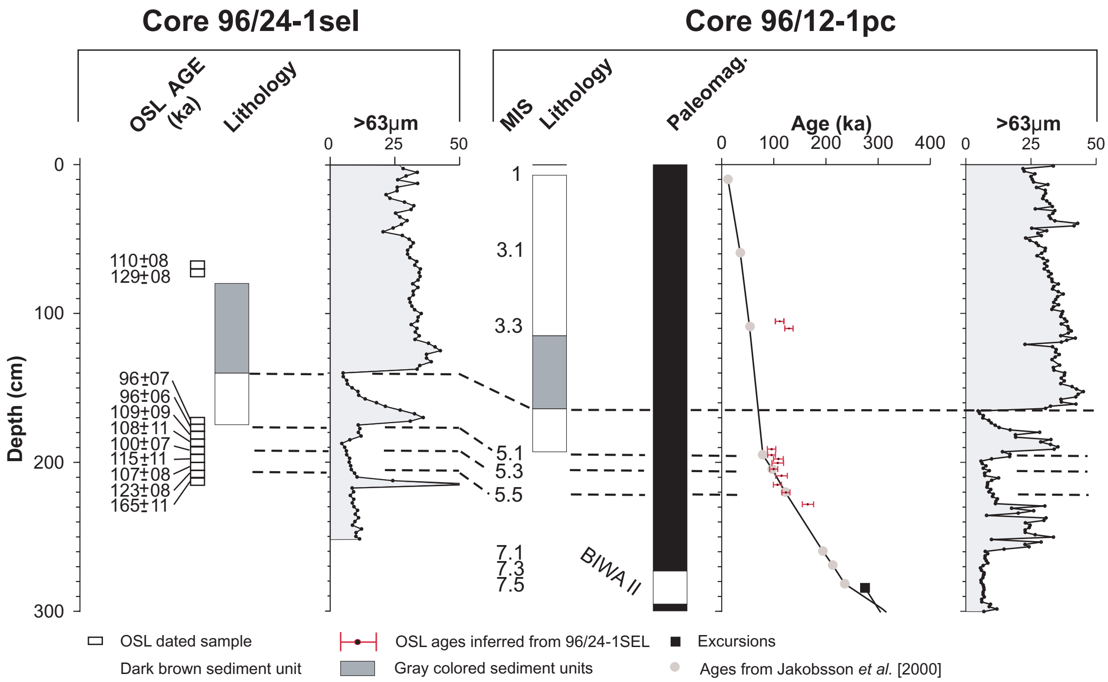

Figure 3. Result from OSL dating of core 96/24-1SEL and the correlation to core 96/12-1PC. Only the uppermost 3 m of the two cores are shown. See Figure 1 for core locations.

clear that this does not compromise the D_e determination in these samples.

[13] The luminescence age (Table 1) is obtained by dividing the estimate of D_e by the dose rate.

4. Results

[14] The OSL dating results of core 96/24-1sel and its correlation to core 96/12-1pc [*Jakobsson et al.*, 2000, 2001], using the >63μm fraction and sediment lithology, are summarized in Figure 3. Each OSL sample is 5 cm thick, and thus represents the integrated age over a 5 cm long interval. Through core correlation of grain size and sediment color (Figure 3), the OSL ages are inferred in the stratigraphy of core 96/12-1pc. The OSL ages thus can be compared to the age model of *Jakobsson et al.* [2000]. In general, the OSL ages of samples from the section of core 96/24-1sel, corresponding to the

Table 1. Summary of Optically Stimulated Luminescence (OSL) Age Data

Lab. No.	Depth, cm	Age, ka	Equivalent Dose, Gy	(n)	Dose Rate[a], Gy ka^{-1}	W.C., %
006634	65–70	110 ± 8	273 ± 8	20	2.48 ± 0.15	32
006633	70–75	129 ± 8	313 ± 4	32	2.44 ± 0.14	28
006613	170–175	96 ± 7	254 ± 7	20	2.64 ± 0.15	31
006612	175–180	96 ± 6	294 ± 8	25	3.07 ± 0.16	38
006611	180–185	109 ± 9	278 ± 20	26	2.55 ± 0.10	46
006610	185–190	108 ± 11	270 ± 25	27	2.44 ± 0.10	39
006609	190–195	100 ± 7	270 ± 15	25	2.68 ± 0.10	43
006608	195–200	115 ± 11	288 ± 24	37	2.59 ± 0.10	43
006607	200–205	107 ± 8	261 ± 8	19	2.42 ± 0.14	43
006606	205–210	123 ± 8	317 ± 6	33	2.59 ± 0.15	43
006605	210–215	165 ± 11	307 ± 8	21	1.86 ± 0.09	42

Radionuclide concentrations converted to dose rates using the data of *Olley et al.* [1998]. W.C. stands for water content defined as weight of water in sample/weight of dry sample.

[a] Cosmic ray dose rate negligibly small. An internal quartz dose rate of 0.06 Gy ka^{-1} has been included in the total.

MIS 5.5 through MIS 5.3 interval in core 96/12-1pc, are in agreement with *Jakobsson et al.*'s [2000] chronology and the samples from above MIS 5.3 to MIS 5.1 are all indicating slightly older ages.

[15] The lowermost of all the samples was taken just below a 5 cm dark brown unit that correlates to a dark brown unit in core 96/12-1pc, a unit that has been assigned to MIS 5.5 [*Jakobsson et al.*, 2000]. An age of 165 ± 11 ka indicates that this sample contains sediments of pre-MIS 5.5 deposition, in agreement with its stratigraphic position below the dark brown unit in core 96/24-1sel (Figure 3). The next following sample, at 205–210 cm, yields an OSL age of 123 ± 8 ka. This sample was retrieved in the brown unit assigned to MIS 5.5, and is in good agreement with the expected age (midpoint MIS 5.5: 122 ka, *Bassinot et al.*, 1994).

[16] The following sample, at 200–205 cm, is dated to 107 ± 8 ka, which is within the inherent OSL estimated uncertainty confirming the chronology proposed by *Jakobsson et al.* [2000]. Further upcore, (195–200 cm) the dated sample includes a portion of the dark brown unit interpreted as MIS 5.3 according to the chronology of 96/12-1pc. The OSL age of 115 ± 9 ka is also consistent with the chronology of core 96/12-1pc to within two standard errors. The remaining part of the brown unit, assumed to represent MIS 5.3, was sampled and OSL dated to 100 ± 7 ka in excellent agreement with the expected age of MIS 5.3 [97 ka; *Bassinot et al.*, 1994]. The four uppermost samples (175–190 cm), in the lower suite of nine OSL dated samples (Figure 3), cover the interval from the top of the brown MIS 5.3 unit to the top of the next following dark brown unit that correlates to MIS 5.1 in core 96/12-1pc. The four OSL estimated ages of these samples yield ages older than the chronology by *Jakobsson et al.* [2000].

[17] The uppermost OSL sample pair, at 65–70 cm and 70–75 cm, was analyzed on samples located in a coarse grain interval (ca 38% >63 μm), within glacial stage MIS 3, and yielded ages of 129 ± 8 and 110 ± 8 ka, respectively. Both these age estimates are at least twice as old as Bassinot's tuned estimates of the MIS 3 interval (24–57 ka). These estimates clearly support the working hypothesis that OSL dated samples representing glacial intervals should yield random old ages caused by incomplete initial zeroing of the particles.

[18] We conclude that the OSL results fail to reject our null hypothesis that cm-scale/ka sedimentation rates persisted during the Pleistocene in the central Arctic Ocean.

5. Discussion and Conclusion

[19] The results of OSL dating support the chronology of *Jakobsson et al.* [2000] for core 96/12-1pc. This implies that cm-scale/ka rather than mm-scale/ka sedimentation rates prevailed during the Pleistocene in the central Arctic Ocean, and that the first negative polarity interval in published Lomonosov Ridge cores does most probably not represent the Brunhes/Matuyama boundary at 780 ka, but more likely the Biwa II excursion presently dated to 275 ka [*Langereis et al.*, 1997; see also *Gubbins*, 1999]. The uppermost reversed polarity interval in cores retrieved from the Amerasian Basin traditionally has been interpreted as the Brunhes/Matuyama boundary and served as a backbone for sediment core chronologies [e.g., *Clark*, 1970; *Witte and Kent*, 1988; *Phillips and Grantz*, 1997]. Several studies have shown that short cores raised within the Arctic basin can be correlated over wide distances because of their apparent similarities in lithostratigraphic and color properties [*Clark et al.*, 1980; *Thiede et al.*, 1990]. For example, *Morris et al.* [1985] correlated cores between the Lomonosov Ridge and the Alpha Ridge. We suggest that the polarity stratigraphy of core 96/12-1pc from the Lomonosov Ridge can be correlated to several Amerasian Basin cores, implying that the uppermost negative polarity interval in those cores represents an excursion within the Brunhes Chron rather than the Brunhes/Matuyama boundary. This interpretation is supported by the recent paleointensity study by *Nowaczyk et al.* [2001] of cores from the Makarov Basin, in which several excursion events are indicated for the Brunhes Chron.

[20] It follows that late Pleistocene sedimentation rates must be increased by a factor that easily can vary between three and ten, depending on whether

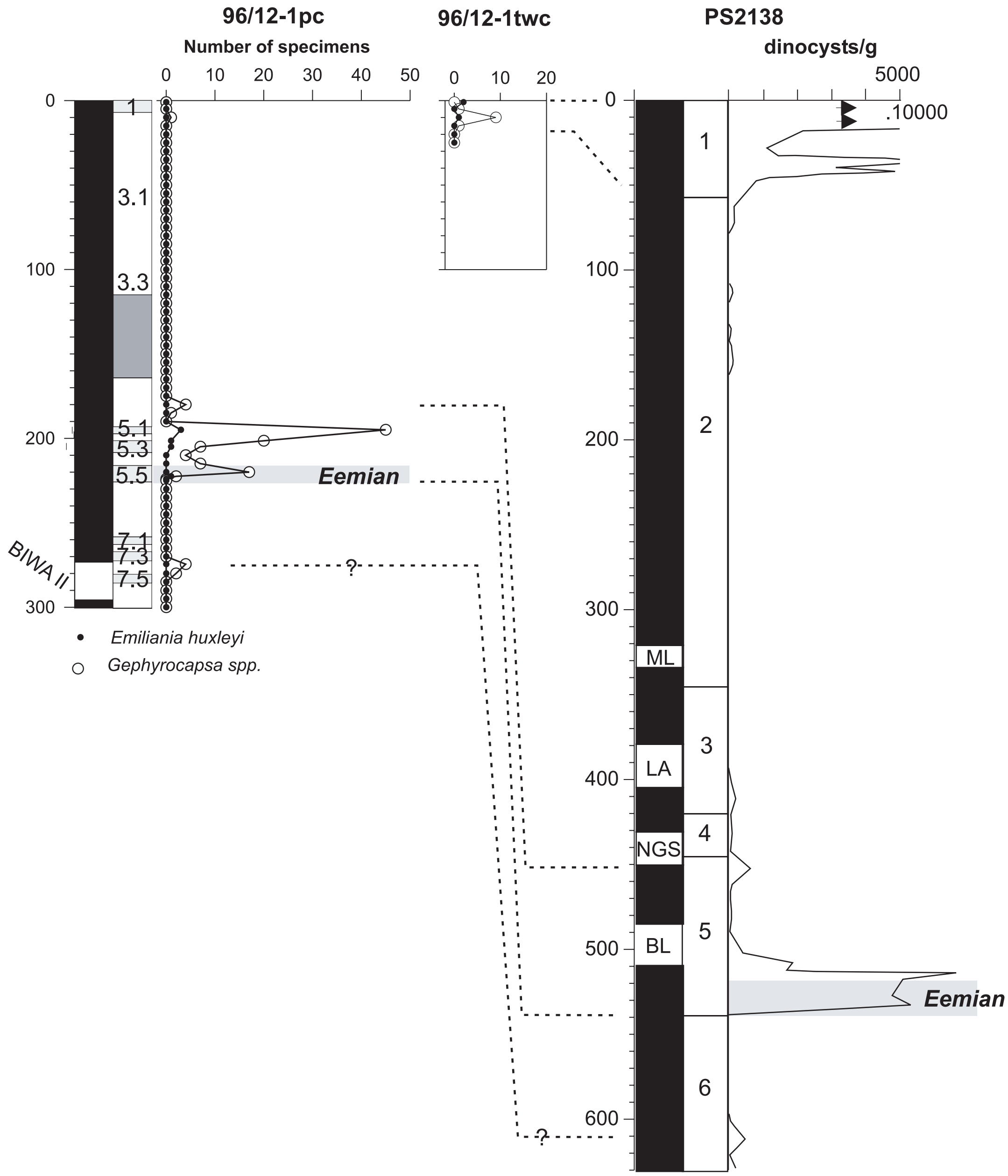

Figure 4. Correlation between core 96/12-1PC from the Lomonosov Ridge and core PS2138 [*Mathiessen et al.*, 2001] from the Barents Sea northern margin based on fluctuations of nannofossil abundance and dinoflagellate taxa.

the first downhole negative inclination represents the Biwa II excursion at 275 ka [*Jakobsson et al.*, 2000], the Lachamp excursion at 40 ka [*Schneider et al.*, 1996], or the Lake Mungo excursion (also referred to as Mono Lake) at about 24–31 ka (*Nowaczyk and Knies* [2000]; L. Benson et al., Age of the Mono Lake excursion and associated tephras, submitted to *Quaternary Science Reviews*, 2002), rather than the Brunhes/Matuyama boundary at 780 ka. The reason why excursions younger than Biwa II (e.g., Lake Mungo, Lashamp, Norwegian-Greenland Sea and Blake) do not consistently appear in the central Arctic Ocean sediments is not clear but factors controlling the preservation of the paleomagnetic signal, such as variations in sediment accumulation, intermittent bioturbation, or erasing of the paleomagnetic signal by postdepositional realignment of magnetic grains, are likely causes. The section above the inferred Biwa II excursion in core 96/12-1pc is characterized by a much higher content of course grained material (Figure 3), which might affect the resistance for postdepositional realignment.

[21] The revised interpretation of the polarity stratigraphy in Arctic cores is essential for interpretations of the paleoenvironmental development in the Arctic Ocean. For example, *Spielhagen et al.* [1997] showed, using their early age model, that enhanced ice rafting was imprinted in Lomonosov Ridge sediments at ca 700 ka in the central Arctic Ocean. They used lithostratigraphic data from cores PS2185-3 and PS2185-6 to suggest that a large northern Siberian ice sheet extended across the continental shelf at ca 700 ka. Applying the younger timescale through core correlation between PS2185-6 and 96/12-1pc, the initiation of the "700 ka" event with enhanced ice rafting, must have occurred well over half a million years later, during MIS 6 at ca 150 ka [*Jakobsson et al.*, 2000, 2001]. This conclusion is currently also supported by Spielhagen who recently revised the age model for PS2185-6, resulting in cm-scale/ka sedimentation rates (Autoref: EUG Meeting in Strasbourg, France, 2001). Moreover, the recent observation of ice grounding at 1 km water depth on the Lomonosov Ridge [*Jakobsson*, 1999; *Polyak et al.*, 2001] has been dated, using the revised stratigraphy, to MIS 6 [*Jakobsson et al.*, 2001]. This implies that the initiation of large-scale glaciation of the Siberian shelf areas (Barents and Kara Seas) responsible for the IRD peaks in the Lomonosov Ridge cores occurred during MIS 6, and that this glaciation created the >850 m thick ice that caused the ice grounding at 1 km present water depth in the central Arctic Ocean.

[22] Recently published results from well-dated cores located at the northern Eurasian margin show fluctuations in concentration of dinoflagellate taxa suggested to be linked to variable inflow of relatively warm Atlantic water into the Arctic Ocean [*Mathiessen et al.*, 2001]. In analogy, intervals with nannofossil abundance peaks in studied sediment cores from the Arctic Ocean have been linked to inflow of north Atlantic water during times that coincide with interglacial periods [*Gard*, 1993]. This forms the basis for correlation between the nannofossil abundance peaks in core 96/12-pc and the variations in concentration of dinoflagellate taxa in core PS2138 located northeast of Svalbard studied by *Mathiessen et al.* [2002] (Figure 4). The events correlate during MIS 5 and the Holocene and, thus corroborates the OSL supported chronology of core 96/12-1pc. This implies that times of interglacial warming and decreased ice coverage have repeatedly brought surface living plankton into the central Arctic Ocean, as noted by *Gard* [1993] for the Holocene. Finally, the OSL dates support the hypothesis that sediment transport in the Arctic Ocean chiefly occurs via icebergs during glacials and via sea-ice during interglacials.

Acknowledgments

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